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Abstract

A multiple wavelength radiometer instrumentation has been developed and used to measure solar irradiance in a water vapor absorption band in the near-infrared region. From these measurements the total atmospheric optical depth at each wavelength of observation is deduced using a linear least-squares fitting method. An iteration technique, based on a power law wavelength dependence of aerosol optical depth, is employed to retrieve aerosol optical depth from the total optical depth data. The technique permits simultaneous determination of precipitable water vapor amount.

Introduction

Information concerning the properties of atmospheric aerosols has been obtained from solar transmission measurements in the ultra-violet, visible, and near-infrared regions. King and Byrne (1976) used solar irradiance data, measured at varying solar zenith angles, to obtain both aerosol optical depth and total ozone content. More recently, Flittner et al. (1993) proposed a technique for inferring total ozone and aerosol optical depths using similar radiometric data. Several researchers (Yamamoto and Tanaka, 1969; Quenzel, 1970; Grassl, 1971; and King et al., 1978) have also demonstrated that parameters describing aerosol size distribution can be deduced from wavelength dependent aerosol optical depth data. This data is usually obtained from total atmospheric optical depth measurements by making corrections for molecular scattering and absorption due to atmospheric constituents such as water vapor, carbon dioxide and ozone. Whereas the contribution due to scattering can be easily calculated, molecular absorption can only be avoided if spectral measurements are made at wavelengths where there is no molecular absorption. However, from the ultra-violet to the near-infrared region there are spectrally broad absorption bands. Two such broadband absorption regions are the Chappuis band due to ozone, ranging from 440 to 800 nm, and the water vapor absorption band ranging from about 880 to 980 nm. Taking measurements outside these regions may lead to the loss of significant spectral information. In this work, measurements were made in five spectral channels between 700 and 975 nm. Three of the channels were located in the water vapor absorption region. A method is described for extracting both aerosol optical depth and precipitable water vapor amount.

Materials and Method

Determination of Aerosol Optical Depth.--The directly transmitted solar irradiance may be given by Beer-Lambert law as

$$F(\lambda) = F_0(\lambda) \exp[-\tau_t(\lambda) m(\theta_0)] \quad (1)$$

where $F(\lambda)$ is the solar irradiance at wavelength λ reaching a ground-based detector; $F_0(\lambda)$ the irradiance incident on top of the atmosphere; $m(\theta_0)$ the atmospheric air mass, a function of the solar zenith angle θ_0; and $\tau_t(\lambda)$ the total optical depth. Air mass is defined as the ratio of the slant path length for the solar rays through the atmosphere to the path length if the sun were in the zenith. This ratio is usually equal to the secant of the solar zenith angel, except for large angles when atmospheric refraction and earth curvature need to be taken into account. Taking the natural logarithm of (1) yields

$$\ln F(\lambda) = \ln F_0(\lambda) - \tau_t(\lambda) m(\theta_0) \quad (2)$$

A plot of $\ln F(\lambda)$ versus $m(\theta_0)$, known as Langley plot, yields a straight line with negative slope $\tau_t(\lambda)$, and y-intercept of $\ln F_0(\lambda)$, assuming the optical depth remains constant during the course of measurement. Thus, using irradiance measurements made at varying air mass, the Langley plot may be used to obtain the total optical depth, $\tau_t(\lambda_i)$, at each wavelength of observation. From the values of total optical depth, τ_t, corresponding values of aerosol optical depth, τ_a, are determined by subtracting contributions due to molecular (Raleigh) scattering, τ_R, and water vapor absorption, τ_W. Thus

$$\tau_a = \tau_t - \tau_R - \tau_W \quad (3)$$

An expression for τ_R, given by Hansen and Travis (1974) for standard surface pressure, P_0 = 1013.25 mbar, is

$$\tau_R(\lambda,P_0) = 0.008569\lambda^{-4}(1 + 0.0113\lambda^{-2} + 0.0013\lambda^{-4}) \quad (4)$$

where λ is the wavelength in μm. At pressure P, τ_R may be expressed as

$$\tau_R(\lambda,P) = \frac{P}{P_0}\tau_R(\lambda,P_0) \quad (5)$$

The water vapor optical depth may be given by

$$\tau_W = \zeta \cdot \alpha(\lambda) \quad (6)$$

where $\alpha(\lambda)$ is the water vapor absorption coefficient (per cm) at wavelength λ, and ζ the water vapor amount (cm). Substituting (5) and (6) in (3) gives

$$\tau_a(\lambda,P,\zeta) = \tau_t(\lambda) - \tau_R(\lambda,P) - \zeta.\alpha(\lambda) \quad (7)$$

Since ζ is unknown, τ_a cannot be calculated directly from (7). Instead, an iteration technique, similar to the method of King and Byrne (1976) is used to retrieve the mean water vapor density, ζ_0. The method described here is based on the Ångström power law which expresses the wavelength dependence of aerosol optical depth as

$$\tau_a = \beta\lambda^{-\upsilon} \quad (8)$$

where β and υ are constants. As the unknown parameter, ζ, takes on values from zero to a maximum value, ζ_{max}, aerosol optical depths calculated from (7) are fitted to the model (8) in order to obtain β and υ, as well as the statistic given by

$$\chi^2 = \sum_i \frac{1}{\sigma_i^2}[\tau_a(\lambda_i,P,\zeta) - \beta\lambda_i^{-\upsilon}]^2 \quad (9)$$

where the summation is made over the wavelengths of observation, λ_i, and σ_i represent the standard deviations of $\tau_a(\lambda_i, P, \zeta)$ values. Also, the partial derivative of χ^2 with respect to ζ is determined as

$$\frac{\partial\chi^2}{\partial\zeta} = -2\sum_i \frac{1}{\sigma_i^2}[\tau_a(\lambda_i,P,\zeta) - \beta\lambda_i^{-\upsilon}] \cdot \alpha(\lambda_i) \quad (10)$$

When χ^2 attains a minimum, its partial derivative with respect to ζ is zero, and the corresponding water vapor amount, ζ_0, represents the mean value for the period of measurement. Having obtained the constants β and υ, aerosol optical depths may be computed from (8).

Instrumentation.--The radiometer instrumentation comprises a telescope, a motorized filter wheel and controller, an enclosed variable frequency optical chopper and controller, an infrared detector, and a digital lock-in amplifier. The schematic diagram is shown in Fig. 1.

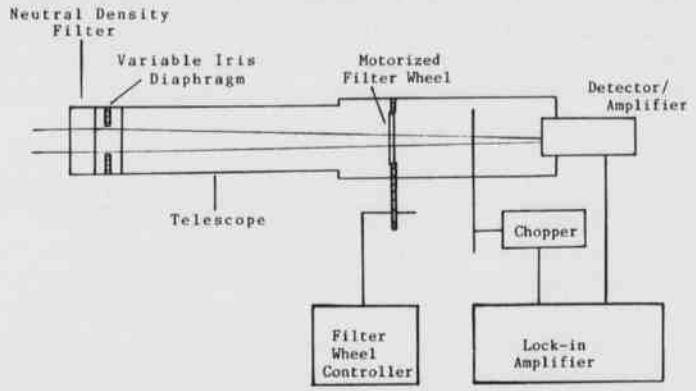

Fig. 1. Schematic diagram of the radiometer instrumentation.

The filter wheel holds five 25-mm diameter interference filters and permits automated wavelength selection. The telescope comprises a 20-cm tube, a neutral density filter, and an aperture stop. The aperture stop is a variable iris diaphragm used to set the field of view to about 3.5°. A neutral density filter is placed at the entrance of the telescope to reduce the intensity of solar flux to a measurable level for the detector. The detector is a radiometrically calibrated silicon photodiode with the National Institute of Standards and Technology (NIST) traceable calibrated responsivity factors for the 400 to 1100 nm range. The digital lock-in amplifier is model 70100 from Oriel Corporation.

Results and Discussion

Solar irradiance measurements at varying solar zenith angles were made on February 16, 1994 at University of Arkansas campus at Pine Bluff (latitude 34.13°N, longitude 92.0°W). Measurements were made with interference filters centered at 700, 875, 903, 941, and 975 nm, respectively. The full bandwidth at half-peak transmittance of each filter is less than 11.5 nm. Readings were taken at approximately 20-minute intervals from 8:00 to 17:00 local time (CST). Solar zenith angels were computed using well-known astronomical formulas and tabulated values taken from the *Astronomical Almanac for the year 1994* (U.S. Naval Observatory, 1993). Fig. 2 illustrates the variation of air mass with local time for the day. Fig. 3 indicates the plot of the natural logarithm of the measured solar irradiance, $\ln F(\lambda)$, versus air mass, $m(\theta_0)$, for the five spectral channels. The lines are the best least-square fit to the data while the resulting slopes give the total atmospheric optical depths for the respective wavelengths. The iteration procedure described above is performed with ζ given values from zero to ζ_{max} in order to determine the constants β and υ, as well as χ^2 and $\partial\chi^2/\partial\zeta$.

The value of ζ_{max} may be deduced from (7) as the maximum value of ζ which produces zero aerosol optical depth. In the region of measurement, the strongest water vapor absorption occurs at about 941 nm, thus

$$\zeta_{max} = \frac{\tau_t(941) - \tau_R(941)}{\alpha(941)} \qquad (11)$$

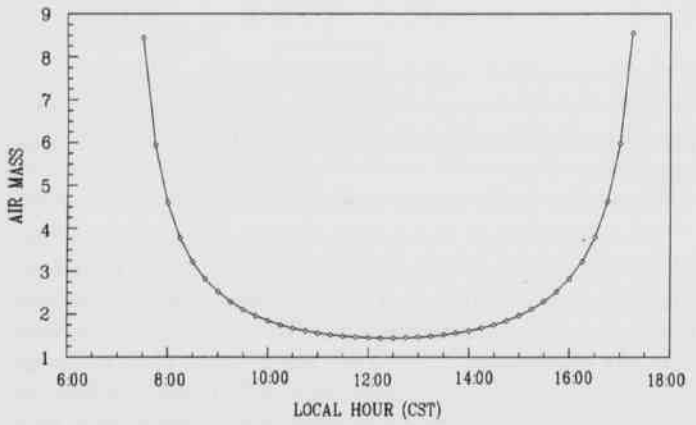

Fig. 2. Variation of atmospheric air mass with local time at Pine Bluff on February 16, 1994.

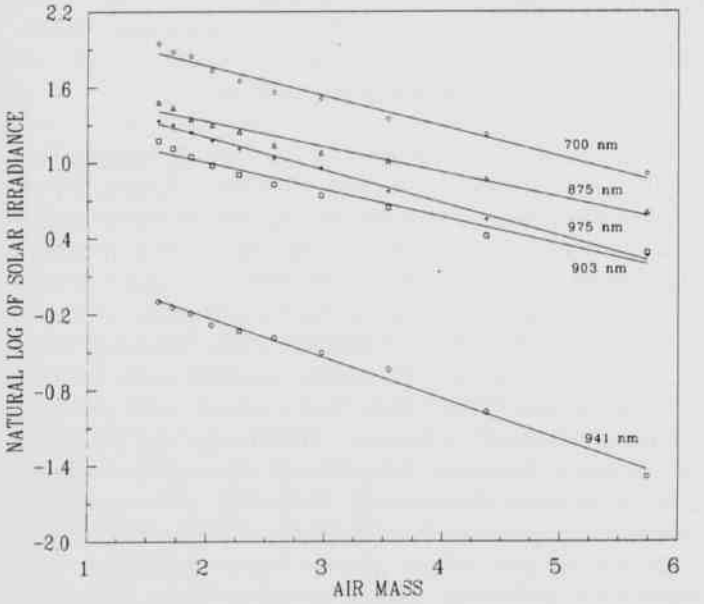

Fig. 3. The natural logarithm of solar irradiance versus atmospheric air mass.

Water vapor absorption coefficients used in this algorithm were taken from the revised Neckel and Labs spectrum in Bird (1984). Figure 4 shows the variation of χ^2 and $\partial\chi^2/\partial\zeta$ with water vapor amount, ζ. The mean water vapor amount, ζ_0, was determined to be 2.2 mm while β = 0.186 and υ = 0.225. The resulting values of aerosol optical depth are shown in Fig. 5.

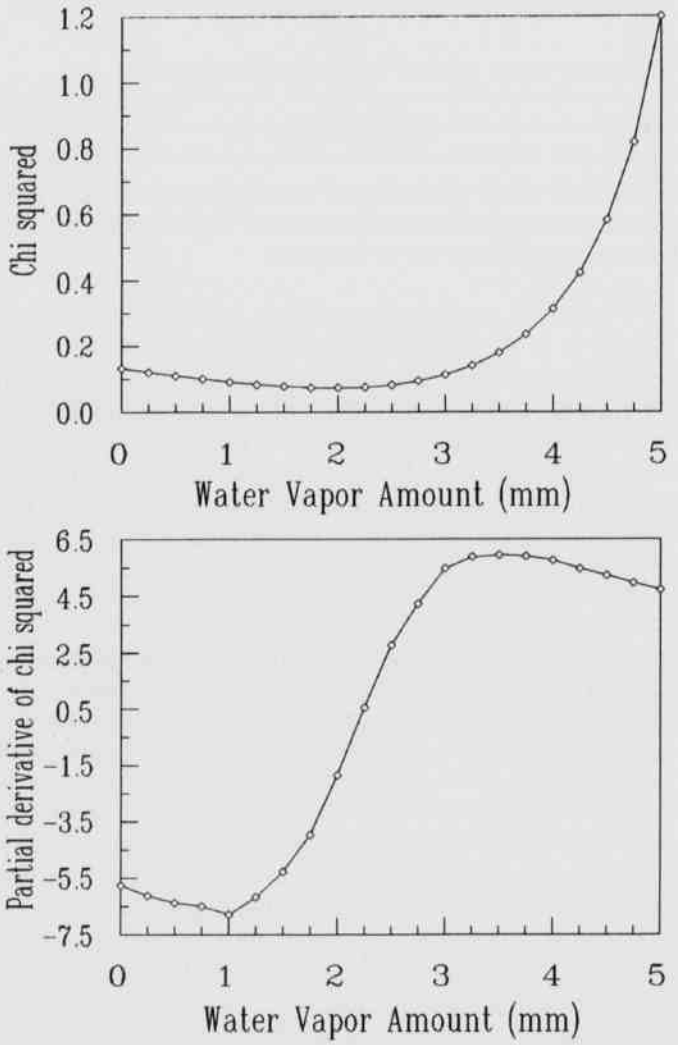

Fig. 4. The relationship of chi-squared statistic and its partial derivative with water vapor amount.

The mean columnar water vapor amount obtained from this algorithm was verified using the well-known differential absorption technique as described in Reagan et al. (1992) and Thome et al. (1992). Here, the two radiometric channels utilized were the 941 nm channel inside the water vapor absorption region and the 875 nm channel outside the absorption region. The ratios of the two radiometric channel measurements were used in a modi-

fied Langley plot. The water vapor amount was retrieved using the transmittance model developed by Thomason (1985). The retrieved value agrees, within an error of 10%, with the amount obtained by the algorithm described above.

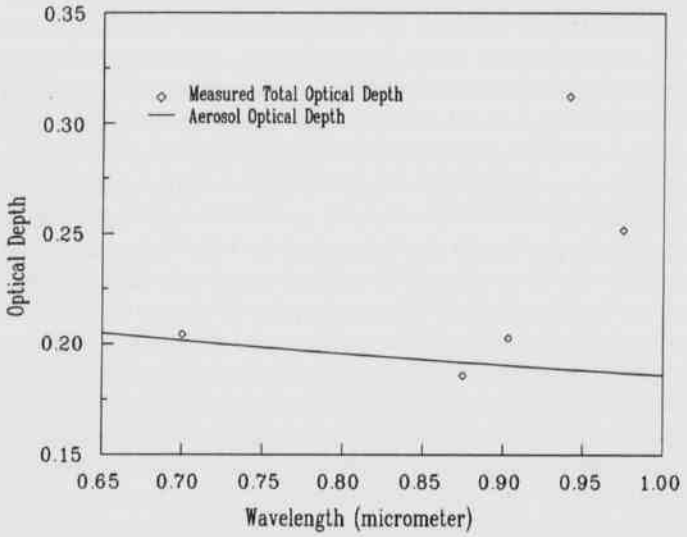

Fig. 5. Atmospheric optical depths measured at Pine Bluff.

Conclusions

The application of the algorithm described in this paper permits the retrieval of aerosol optical depths at wavelengths within water vapor absorption band in the near infrared region. The mean precipitable water vapor amount is obtained through nonlinear least squares fitting by minimizing the χ^2 statistic. The principal assumption upon which the technique is based is that aerosol optical depths can be adequately described by the Ångstöm power law within a relatively narrow wavelength region.

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Literature Cited

Bird, R.E. 1984. A simple, solar spectral model for direct-normal and diffuse horizontal irradiance. Solar Energy 32:461-471.

Flittner, D.E., B.M. Herman, K.J. Thome, J.M. Simpson and **J.A. Reagan.** 1993. Total ozone and aerosol optical depths inferred from radiometric measurements in the Chappuis absorption band. J. Atmos. Sci. 50:1113-1121.

Grassl, H. 1971. Determination of aerosol size distributions from spectral attenuation measurements. Appl. Opt. 10:2534-2538.

Hansen, J.E. and **L.D. Travis.** 1974. Light scattering in planetary atmosphere. Space Sci. Rev. 16:527-610.

King, M.D. and **D.M. Byrne.** 1976. A method for inferring total ozone content from spectral variation of total optical depth obtained with a solar radiometer. J. Atmos. Sci. 33:2242-2251.

King, M.D., D.M. Byrne, B.M. Herman and **J.A. Reagan.** 1978. Aerosol size distributions obtained by inversion of spectral optical depth measurements. J. Atmos. Sci. 35:2153-2167.

Quenzel, H. 1970. Determination of size distribution of atmospheric aerosol particles from spectral solar radiation measurements. J. Geophys. Res. 75:2915-2921.

Reagan, J.A., K.J. Thome and **B.M. Herman.** 1992. A simple instrument and technique for measuring columnar water vapor via near-IR differential solar transmission measurements. IEEE Trans. Geos. Remote Sensing. 30:825-831.

Thomason, L.W. 1985. Extinction of near infrared solar radiation as a means for remote determination of atmospheric water vapor. Ph.D. dissertation, University of Arizona, 125 pp.

Thome, K.J., B.M. Herman and **J.A. Reagan.** 1992. Determination of precipitable water vapor from solar transmission. J. Appl. Meteor. 31:157-167.

Yamamoto, G. and **M. Tanaka.** 1969. Determination of aerosol size distribution from spectral attenuation measurements. Appl. Opt. 8:447-453.

U.S. Naval Observatory. 1993. The Astronomical Almanac for the year 1994. U.S. Govt. Printing Office, Washington, D.C., 539 pp.